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DISCUSSION

PREDICTION OF HORIZONTAL RESPONSE SPECTRA IN EUROPE[1]

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The discussed paper[1] presents a scaling model for Absolute Acceleration Spectra in Europe, to be used in seismic hazard analyses. The discusser will comment on the following aspects of this paper: (1) uniformity of the data set used, (2) applied data processing techniques, (3) consistency in classification of the local site conditions, (4) adequacy of local soil classification used, (5) magnitude scale used, (6) assumed magnitude dependence, (7) some trends in the distribution of the residuals for different types of sites, (8) agreement with results by other authors for a region in Europe (territory of former Yugoslavia) and (9) some aspects related to use of scaling laws for response spectra for seismic hazard calculations (e.g. need for a wider period range of the spectra and for analytic approximation of the distribution of the residuals).

(1) The authors state that they use 'large and uniform data set' (42 per cent of their data is from Italy, 19 per cent from Greece, 13 per cent from the former Soviet Union, 12 per cent from the former Yugoslavia, and the remaining 14 per cent from Algeria, Azores, Bulgaria, Iran, Portugal and Turkey). They use a total of 422 records from 157 earthquakes. Trifunac and Todorovska[2] and Trifunac *et al.*[3] described the differences in attenuation of intensity scales for natural seismological zones of the Balkan Peninsula (Southeast Europe), proposed by Shebalin.[4] The observed differences for these regions are significant, but as yet not fully researched and understood. Thus, it is not likely that attenuation of strong motion in Italy, Greece, former Yugoslavia, and former Soviet Union can be described by one common law. Furthermore, it is difficult to describe the attenuation of strong motion by only a frequency-dependent exponent of source to station distance. Other important factors which influence the attenuation function are e.g. the 'size' of the seismogenic zone (distributions of fault width, length and depth and orientation of the fault plane) and the stress in the source region, which could be modelled via the source dimension S and the coherence length S_0.[5,6] The attenuation law of the form r^n, where n is a frequency-dependent exponent, is inadequate for small and for large r (e.g. see Figures 6 and 7 in the discussed paper[1]) and cannot be used to describe near-field, intermediate field, far-field and surface wave attenuation, all by a single value of n.[7,8] Thus, the data used by Ambraseys *et al.*[1] is not uniform with respect to the different attenuation laws which apply in the regions contributing the data. Strong motion data recorded in former Yugoslavia (325 three-component records from 183 earthquakes) has been used to develop a frequency dependent attenuation function[7] and scaling laws of response spectra[9] specific for that region. Correlation of peak amplitudes of strong motion with one of the intensity scales for that region (Mercalli–Cancani–Sieberg scale)[10] and frequency-dependent duration of strong ground motion[11,12] have also been studied using the same database. All of these studies[7–12] discuss the differences between attenuation of strong ground motion in California and in former Yugoslavia. Some of the differences have been interpreted in terms of different soil and geologic settings of the recording stations.

(2) The acceleration data used by the authors should have been corrected for the instrument response (using nominal values for the instrument constants). The natural frequency of AR-240 accelerographs is

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$f_N \sim 18$ Hz. Absolute acceleration spectral amplitudes are not affected significantly by the instrument transfer function for frequencies $f \ll f_N$, but the slope of the coefficients C_1' and C_2 and C_4 versus period is affected near period $T = 0{\cdot}1$ s. This may prove important for other studies[13] which may depend on the authors calculations. Elliptical filters for baseline correction of strong motion accelerograms require time reversal of the record. This is not acceptable[14,15] and can be avoided by employing different filters.[16]

(3) The authors are inconsistent in referring to the local soil site conditions. The terms 'soil conditions' and 'site geology', both seem to refer only to the local soil conditions. This is next complicated by renaming categories A, B, C and D, as defined by Boore *et al.*,[17] to respective categories R, A, S and L. This becomes even more confusing (erroneous) when the authors use (associate) terms as 'rock' and 'stiff soil' with their categories R and A. Those terms, as proposed and used by Seed[18] and later adopted by Trifunac,[19] belong to a different type of classification from the A, B, C and D classification (based on average shear wave velocity in the top 30 m).[17] Groups R, A and S thus do not classify the sites by any size (depth) of the soil deposits, but only by the average shear wave velocity near the surface. The Seed's soil site classification,[18] termed $s_L = 0$, 1 and 2, has the depth parameter built into its definition. 'Rock' ($s_L = 0$), 'stiff soil' ($s_L = 1$) and 'deep soil' ($s_L = 2$) characterize, respectively, soils with shear wave velocity less than 800 m/s and depth less than about 10 m, shear wave velocity less than 800 m/s and depth less than 75–100 m, and shear wave velocity less than 800 m/s and depth from about 100 to 200 m. At the other end of scale and different from the A, B, C and D, and from the Seed's local soil classification is the site geology classification,[20] introduced in 1975 in terms of parameters $s = 0$ (sediments), $s = 1$ (intermediate) and $s = 2$ (basement rock), and later refined to include the depth of sediments.[21] The geologic classification describes the local site conditions on a different scale from the soil classifications. Both classifications must be considered simultaneously in the characterization of site specific spectra, as ignoring the local geologic conditions may lead to exaggerated factors 'representing' the local soil conditions.[22] Until recently, analyses characterizing spectral amplitudes have been limited mainly to the above-mentioned characterizations of the local site conditions. 'Horizontal' measures of geological site characteristics,[23] 'type' of propagation path and percentage of sediments along the propagation path[24] and their influence on peaks of recorded motion and spectral amplitudes were introduced in 1995. Systematic gathering of many of these site parameters is very expensive and time consuming, but the geological site description in terms of $s = 0$, 1 and 2 is simple and easy to perform.

(4) A recent study of response spectrum amplitudes of strong motion in California,[25] concluded that A, B, C and D local soil classification (based only on average shear wave velocity in the top 30 m) becomes statistically insignificant when used simultaneously with the $s_L = 0$, 1 and 2 and $s = 0$, 1 and 2 classification. This implies that classifications that include also the depth of the soil deposit are better than those based only on information close to the surface. In view of the above, coefficients C_A and C_S in the discussed paper do not describe adequately the effects of the local soil conditions A and S on the amplitudes of absolute acceleration spectra in the region studied.

(5) The purpose of the discussion on seismic moment–magnitude relationship[1] is not clear, because the authors use surface wave magnitude, M_S, only. For strong motion data in California, there is also a change in slope in the linear relationship of the logarithm of seismic moment, M_0, versus magnitude, M, near $M = 5{\cdot}5$. In contrast to equation (2) in the discussed paper,[1] the slope of M versus $\log_{10} M_0$ decreases for $M < 5{\cdot}5$.[6] For small earthquakes, the corner frequencies in the source spectrum are underestimated, because of low values of the quality factor Q.[6] If not corrected, this will lead to overestimating the source area and seismic moment, as shown in Figure 2 of the discussed paper. The moment magnitude is indeed becoming a popular parameter in descriptions of attenuation laws, but this should not be so unless it can be shown that this leads to accurate estimates of strong motion amplitudes. The moment magnitude, a teleseismic and long-period estimate of the energy release from the earthquake source, has been inspired by the seemingly straight line correlation of magnitude with $\log_{10} M_0$. A further study of the physical bases in this correlation would show that the 'constant' 10·7 in equation (1) of the discussed paper should depend on the apparent stress in the source region, while the slope $\frac{3}{2}$ depends on the rate of growth of the fault area with M_S.[6] Therefore, these trends are not the same in different tectonic regions (i.e. Balkans, Italy, Turkey, former U.S.S.R, etc.).

For local magnitudes $M_L \lesssim 4$, the surface wave magnitude, M_S, computed in the discussed paper is systematically smaller (by 0·3–0·4 magnitude units) than the reported M_L. For for small events, in

California[26,27] and in Yugoslavia,[28] the local magnitude computed from strong motion data, M_L^{SM}, differs from the local magnitude, M_L, as a result of the rapid attenuation of high-frequency waves with distance, and the difference[29] exceeds one magnitude unit for $M_L \lesssim 5$. Thus, the 'benefits' of working with well-defined teleseismic and long-period estimates of magnitude are at the cost of loosing information on the high-frequency strong motion waves. The type of magnitude used for scaling of response spectra should not be chosen only on the basis of excellent correlation with seismic moment M_0. Distant (teleseismic) estimates of magnitude should be avoided, because the longer propagation paths increase the uncertainties associated with attenuation. Engineering estimates of future strong motion must be as reliable as possible, and must portray realistically all (including high) frequencies of strong motion. This may require careful calibration of the magnitude scales which are used in different areas, and, whenever possible, this calibration should be based as directly as possible on strong motion data recorded close to the earthquake source.[26,30]

(6) Analyses of the physical processes at the earthquake source,[6] as well as numerous empirical studies of peaks and of spectral amplitudes of strong motion,[31–34] show saturation of strong motion amplitudes for magnitudes greater than about 6·5. Therefore, the authors should have considered at least one additional term in their scaling equation, for example $C_5 M_S^2$. Their data contains five earthquakes with $M_S > 7{\cdot}0$, with most recordings in Iran.

(7) The plots of residuals of peak acceleration and of selected spectral amplitudes versus distance (Figures 10, 13 and 16 in the discussed paper) show positive slopes for sites labelled 'R', and smaller (occasionally negative) slopes for those labelled 'A' and 'S'. Such trends have been incorporated into the attenuation equations for strong motion data in California,[24,25] and can be incorporated in scaling laws of spectral amplitudes for selected subsets of European strong motion data.

(8) Figure 1 compares absolute acceleration spectra of the discussed paper[1] (heavy dahsed lines) with pseudospectral acceleration in Former Yugoslavia[9] (shaded intervals). The plot on the left compares results for $s_L = 0$ ('rock')[9] with those for R sites,[1] and the plot on the right, results for $s_L = 1$ (stiff soil)[9] with those for A sites,[1] at epicentral distance equal to 30 km. To facilitate comparison of these two different approaches and to reduce the observed differences, M_S of Ambraseys *et al.*[1] was 'corrected' by using the approximation

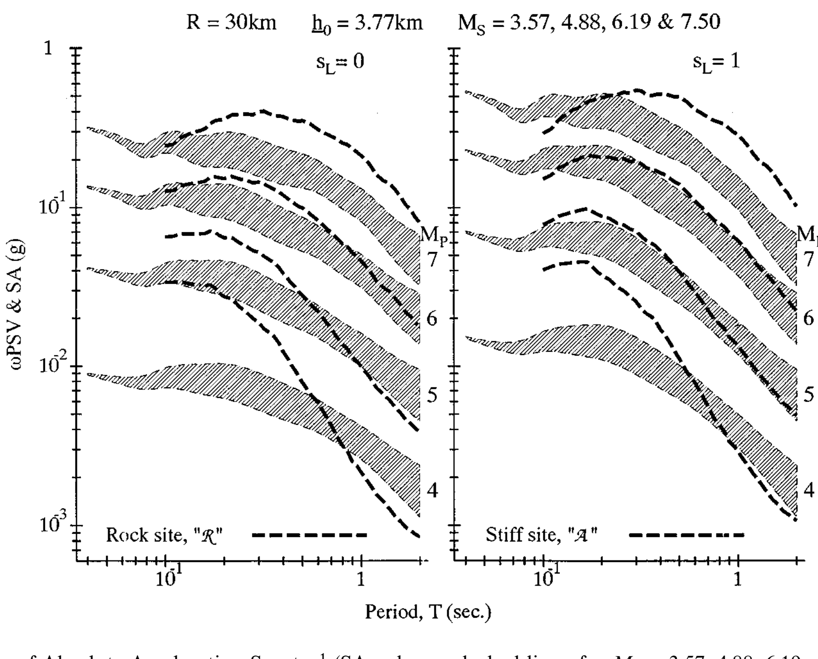

Figure 1. Comparison of Absolute Acceleration Spectra[1] (SA-g, heavy dashed lines, for $M_S = 3{\cdot}57$, 4·88, 6·19 and 7·50) with pseudo-acceleration spectrum (ω PSV-g) in former Yugoslavia (shaded zones for M = 4, 5, 6, and 7) for $R = 30$ km. Top boundary of dashed zones corresponds to $s = 0$ (sediments) and bottom to basement rock ($s = 2$). Left figure is for $s_L = 0$ ('rock') soil conditions, and right for $s_L = 1$ 'stiff soil sites'. Those are compared with spectra computed for rock sites '$\mathscr{R}$' (left) and soil sites '$\mathscr{A}$' (right)[1]

$M_S \sim -1{\cdot}67 + 1{\cdot}31 M_L$, so that $M_S \sim 3{\cdot}57$, $4{\cdot}88$, $6{\cdot}19$ and $7{\cdot}50$ corresponds roughly to $M = 4$, 5, 6 and 7 (Figure 1). The top and bottom boundaries of the gray zones correspond, respectively, to geologic conditions $s = 0$ (sediments) and $s = 2$ (basement rock). The observed discrepancies between these two scaling models are so large that one should understand and account for all relevant contributions to these differences before these models are used in any future calculations. At epicentral distances equal to 5 and 100 km, these differences are even larger.

(9) For probabilistic estimates of seismic hazard, reliable estimates of the rate of earthquakes occurrence versus magnitude (intensity) are essential.[35–40] Also, before a scaling model of absolute acceleration spectra can be used in seismic hazard analyses, it is necessary to develop description of the distribution of the residuals relative to the scaling model. The period range of the spectra in the discussed paper, 0·1–2·0 s, may be too narrow for some applications (long period structures, such as water tanks, bridges, tall buildings and base isolated buildings, and for high-frequency equipment). This period range can be extended[6,13,41,42] so that uniform hazard maps of required scaling parameters, can be prepared for a broad period range.

With the above-proposed modifications, the scaling of strong motion amplitudes in Europe could be improved, not only for direct estimation of strong motion for a given earthquake, distance and site conditions, but also for use with various seismic hazard analyses leading to site specific spectra, or to various microzonation maps.

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